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by

P R Rowntree

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THE WATER BUDGET OF MIDDLE LATITUDE CONTINENTAL REGIONS - A MODELLING AND OBSERVATIONAL STUDY

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Abstract
An attempt is made to use ground-based data to validate the modelled surface water budget in a region of Russia where the Hadley Centre climate model becomes too arid in summer. Observed data show significant imbalances, and so are of limited value for use in the budgets needed for validation of models. However, they do indicate that model soil moisture is too low in spring, leading to unrealistic drying out in summer. A likely cause is lack of precision in definition of the soil characteristics, together with uncertainties in derivation of the model's soil parameters; the specification of the root depth may also contribute.

Introduction

An early and important finding from climate models concerning the effects of increasing greenhouse gases was that the middle latitude continents would tend to have lower soil moisture in summer. For example Mitchell et al. (1990) showed widespread decreases in soil moisture in summer with the three high resolution models used for the main assessment. The confidence attached to this prediction in the 1990 IPCC Report was not high (2 stars out of a maximum of 5); one reason for this was that the simulations even of zonally averaged soil moisture differed greatly from each other (Gates et al., 1990). The IPCC Report also included a validation of the soil moisture over part of Russia, for three high resolution models. This validation was against the data of Vinnikov and Yeserkepova (1990) (V&Y hereafter) who summarised seasonal trends in soil moisture, mainly over the period 1972-85, using the large collection of data available in the former USSR. The Meteorological Office Hadley Centre (HC) model was shown to be close to the observations in that region (Gates et al., 1990).

Model

However, since then, the HC model has been considerably revised; though the physics were largely unchanged, a new dynamics scheme was introduced, giving more realistic levels of eddy activity, and the simulations are now substantially different. The versions of the model used here are similar to those described by Jones et al. (1995).

The immediate objective of this paper is to assess the feasibility of using ground-based data for validating the surface water budget of the model, with the

longterm aim of improving the realism of the simulation of the surface hydrology. The realism of the model is compared against the observed for part of Russia, using V&Y's data discussed above. Two versions of the model are considered. The second version (M2) differs from the first (M1) in the following respects:

reductions in the horizontal diffusion of atmospheric specific humidity, potential temperature and horizontal wind,
modifications (decreases) to the evaporation of falling precipitation (correction of the model formulation),
a change in the cloud scheme which tended to increase the cloudiness in summer.

The reduction in diffusion was suggested by earlier reductions in moisture diffusion which were found to improve the simulations especially near mountains, with less intense orographic maxima and increased precipitation over adjacent lowlands. The changes in evaporation of precipitation gave some increases in summer rainfall. The cloud scheme was changed because the scheme in M1 was found to give unrealistically small cloud amounts in short-range predictions; this change was not wholly beneficial; in particular it increased top-of-atmosphere albedos particularly over high latitude continents giving substantially higher values than observed (Hall, 1995).

Experiments

The two experiments (E1 and E2), using respectively M1 and M2, were run for 20 years and for 10 years 3 months respectively, using climatological sea surface temperatures and sea-ice, starting from an analysis made using the MO Unified Model and assimilation system for early June. The initial soil moistures were taken from a simulation using the previous version of the Met Office climate model. The results discussed here are the means for the full duration of the experiments except the first 3 months of E2. Detailed monthly mean data were extracted for gridpoints at 2.5° latitude 3.75° longitude intervals on a section from 60°N 33.75°E to 45°N 56.25°E, through an area of Russia relatively well-covered by V&Y's data and with a significant tendency to summer aridity in the model. The discussion in this paper focusses on the water budget at 55°N 41.25°E, with only limited discussion of the other points.

The surface water budget and availability of observed data

The soil moisture balance may be written:

$$d(SM)/dt = R + M_S - E_W - Y \qquad (1)$$

Here, the terms are the snowmelt (M_S), the rainfall (R), surface runoff and drainage from the soil layer

considered (Y), evaporation of water (E_w) and change in soil water content (SM). All these terms are available from the model. However, for reality, the situation is less favourable. In particular, the data on rainfall, as distinct from snowfall (S), and for snowmelt are not readily available. Outside the snow season however, equ. (1) can be simplified to:

$$d(SM)/dt = P - E - Y \qquad (2)$$

This equation can also be used for the annual mean water balance as between times with no snow cover:

$$S = M_S + (E - E_W)$$

I consider below for each term in turn the data which were readily available for this study.

Soil moisture change (d(SM)/dt): V&Y have published data on soil moisture in the top metre of soil for about 50 Russian soil moisture stations; these are all west of 105°E and south of 60°N. apart from three north of 60°N west of the Urals. Most are for the period 1972-85. Other Russian data have been used by Robock et al. (1995). The data are means over the period for the 28th of most months, though with limited data from snow-covered periods of the year.

Precipitation (P): The precipitation data for 1972-85 were available for a rather sparse network of stations from the archives maintained in the Hadley Centre. These are based mainly on the data received on the GTS (Global Telecommunications System), supplemented where necessary by other sources such as World Weather Records. However, the rainfall stations mostly do not coincide with the soil moisture stations. A further problem is that these data are likely to require correction for errors inherent in the measurement techniques (Groisman et al., 1991). These errors are most serious during snowfall; corrections to the annual mean data can amount to an enhancement of several tens of percent. An alternative is to use corrected data which are available for stations in histogram form, as annual totals and monthly percentages, and also as a mapped analysis of annual totals, in the World Water Balance Atlas (Korzun et al., 1978) for the period 1891-1970.

Surface runoff and drainage (Y): An analysis of river runoff data is available in Korzun et al. (1978). This is in the form of a mapped analysis of annual totals and monthly percentages in histogram form. The histogram data, being for catchments, are not colocated with either the soil moisture or rainfall stations, but do have quite a dense coverage over European Russia. However, as discussed later, the river runoff does not equal the sum of Y and the change in soil moisture, as

there are often intermediate storage reservoirs such as groundwater.

Evaporation (E): Korzun et al. (1978) also map annual evapotranspiration; the seasonal variation is shown in histogram form, but only for a few points (in the area of our interest for Moscow, Helsinki and Charkov (50°N, 36°E)).

Discussion: To calculate a soil moisture budget including seasonal variation, knowledge of three of the four terms in Eq. (2) would suffice. This is only possible with the available data if we use the data from Korzun et al. (1978). However, for validation purposes, we need a balanced annual budget. Korzun et al. do not provide this, having made estimates of precipitation, evaporation and runoff which are to some extent independent - although they do state that the evaporation calculations use the Budyko complex method, in which monthly evapotranspiration is computed from potential evaporation, and that precipitation values may be used in the checking process. The lack of balance is evident from examination of, say, runoff at points where precipitation and evaporation contours intersect. Nevertheless, the data have been used, albeit with caution, on the assumption that the data are as good as are available.

Table 1 shows the budgets extracted for the model section. Note that there are inevitable read-off errors in using the mapped data. Assuming smooth variations between contours, the estimated uncertainties are about 20 mm for E and P, 10mm for Y, in each case 20% of the contour interval. There are substantial discrepancies as expected from the above discussion. All are positive and so could be reduced by using the gauged precipitation data such as that shown in Table 1 for Leningrad (St Petersburg) against the 60°N point and Moscow (55°N). Also shown in Table 1 are estimated values for Moscow itself from the maps in Korzun et al. (1978). The balance is better here, and as the monthly distribution is also available for P and E we will use these data in the comparison of modelled and observed. The gauged precipitation for Moscow for 1972-1985 in the Hadley Centre archives is 699 mm, close to the 700 mm/year shown by the Korzun et al. map, and well above the mean for 1891-1970 mean of less than 600 mm/year from World Weather Records; whether this is a real increase or one due to changes in site or measurement methods is beyond the scope of this paper (but see Groisman et al., 1991). For Leningrad, the mean for the 1972-85 period was 622 mm, again showing a substantial increase from 1891-1970, but well below the mapped value of 800 mm in Korzun et al.

Table 1: Annual soil water budget for 1891-1970 from Korzun et al. (1978) (mm/year) (also gauged precipitation from World Weather Records for 1891-1970)

Point	P	E	Y	P-E-Y	P_{gauge}
60, 33.75	780	400	290	90	556[1]
57.5, 37.5	730	470	220	40	
55, 41.25	705	520	140	45	597[2]
52.5, 45	610	480	105	25	
50, 48.75	370	335	15	20	
47.5, 52.5	250	250	0	0	
55.6, 37.5	700	510	180	10	597[2]

[1]: Leningrad (St. Petersburg) (59.9°N 30.3°E)
[2]: Moscow (55.6°N 37.5°E)

Possible interactions between precipitation and evaporation

For an assessment of the modelled surface water balance against observations to be useful, we require realistic atmospheric forcing. One of the most important components is the precipitation. However, because of the feedback from evaporation through the atmospheric humidity, errors in the precipitation can be a symptom of faults in the surface parametrization. This is most likely to be the case during summer when the actual or modelled evaporation is limited by soil moisture. Figure 1 illustrates the development of errors in modelled precipitation through the summer. Rainfall in June (and July (not shown)) is quite realistic near 50-70°N, while in August it is clearly deficient; the possible association of the problem with land surface feedbacks is indicated by a corresponding decline in the evaporative fraction

$$\mu = (LE/(LE+H))$$

at 57.5°N and 55°N in Experiment E1 (Table 2). For relatively moist conditions where soil moisture is not limiting evaporation, μ has an asymptotic value which is reached in various limiting conditions including that of large values of the atmospheric resistance (r_A) (Rowntree, 1991). This asymptotic value increases with temperature from about 0.41 at 0°C to 0.69 at 20 and 0.79 at 30°C. With typical summer temperatures near the surface of 20°C near 55°N, a value of about 0.7 is associated with soil moistures near and above the level at which they cease to limit evaporation. This problem is less serious for Experiment E2 than for E1 (Table 2), though still marked at the southern point in August.

Table 2: Modelled values of the ratio of latent heat flux to the sum of sensible and latent heat flux for 57.5°N 37.5°E and 55°N 41.25°E from May to August

Experiment	May	June	July	August
57.5°N 37.5°E				
E1	.727	.744	.701	.547
E2	.701	.765	.796	.739
55°N 41.25°E				
E1	.718	.575	.277	.259
E2	.686	.740	.567	.323

Comparison of the modelled soil moisture (Fig. 2) with observed values from V&Y for near the northern point also suggests that E1 is too dry in late summer. This deficiency appears even more serious in similar diagrams for 55°N (shown later - Figure 7) and 50°N (Figure 3) with available soil moisture almost exhausted by August. To investigate this in more detail, the water balance for the 55°N point is studied in the following sections.

Validation of modelled surface fluxes for 55°N

Precipitation: Figure 4 compares the monthly precipitation for the two experiments with the observed for the 55°N point. The precipitation used here is that for Moscow from Korzun et al. (1978). There is little difference (<1%) between the annual precipitation at Moscow and at the 55°N gridpoint. Note that the gauged data, also shown in Fig. 4, do not differ systematically from the Korzun et al. estimates in April to September - i.e. outside the snow season. Figure 4 suggests that E1 is too dry in the annual mean, with deficient snowfall in winter and too little rainfall from late June to September. E2 again gives more realistic figures but is still too dry in July to September. These differences are mostly consistent with the general character of the differences between the two versions expected from the changes in model formulation.

Radiation: A comparably important component of the forcing is the net radiation which provides the energy for the turbulent surface fluxes of heat and moisture. Observed data on this are less plentiful and less reliable than data for downward solar radiation so we compare this with the observed (Fig. 5), using data for Moscow from Budyko (1963). These relatively early observations, averaging 119 W/m^2, are consistent with an annual mean of 114 W/m^2 quoted by Garratt (1994) from more recent data. Figure 5 shows that the observed downward solar radiation lies between the two

model versions from April to June but is otherwise less than either. The largest overestimates are in late summer and can be attributed to the drying of the modelled climate at this time, with consequent underestimate of cloudiness in both versions. E1 is sunnier than E2 throughout the year, consistent with the previously-noted tendency for less cloudiness; annual means are 147 and 124 W/m^2 respectively. In the comparison of GCM and observed data by Garratt (1994), three of the four models reported excessive values for Moscow, ranging from 130 to 146 W/m^2; the fourth was an early version of the Met Office model, which gave an underestimate.

Evaporation: The evaporation has been compared with the data in Korzun et al. (1978) (WWB) to provide some assessment of the model simulations (Fig. 6). E1 agrees with the WWB estimates in spring but falls well below it from late June to September. E2 on the other hand is deficient in spring and, though better than E1, still deficient in summer. This difference between E1 and E2 in spring is consistent with the differences in downward solar radiation (Fig. 5). Although apparently superior in spring, the large decline in evaporation in June in E1, before there is any serious decline in rainfall, suggests that the soil moisture simulation is not realistic. We consider this in the next section. Also shown in Figure 6 are estimates by Budyko (1963) based on a combined heat and water balance approach. The annual total is much closer to the model estimates, but the shape of the seasonal variation is similar to the estimate by Korzun et al., showing no evidence of drying of the soil in late summer on the scale found in the model. The lesser annual total evaporation is likely to be linked through the water balance calculation with the use of lower estimates of the precipitation based on gauge data available around 1960 (see earlier).

Discussion of Henning (1989)'s data: Estimates of radiation and turbulent heat flux and runoff quantities could also be extracted from the maps of the heat balance in Henning (1989). For 55^oN 40^oE, he estimates an annual evaporation of just over 350 mm, lower than Budyko (400 mm) and Korzun et al. (510 mm) and also the model (E1: 380; E2: 405). He maps only selected months; compared with Budyko's data, his figures are higher in July (at 2.7 mm/day), similar in August, and about 40% lower in April and October. He estimates runoff, using a water balance calculation, at about 190 mm/year, a similar figure to Korzun et al. (1978); this suggests he has used a much lower estimate of precipitation (about 550 mm/year), close to the gauged values for Moscow for 1921-50 in WWR.

Soil moisture for 55N

The V&Y data do not include a station close either to Moscow or the 55°N 41.25°E point. The following analysis therefore uses an average of the nearest available data to assess the model. Figure 7 compares the seasonal variation in soil moisture for the experiments at this point with that in the top metre at four soil moisture stations in V&Y (Table 3) which are close to the same latitude belt and reasonably close in longitude. The shape and magnitude of the model variations are generally similar to the observed, but the mean level is much lower as shown earlier for 57.5°N and 50°N; the soil moisture almost vanishes around August, explaining the low evaporation noted in Figure 7.

Table 3: Data on available soil moisture (mm) in the top metre from V&Y used in validation of model at 55°N 41.25°E

Number	Lat	Long	W_f	W		Change	Range
				28 Apr	28 Jul		
7	58N	36E	120	334	242	-92	106
8	56N	28E	175	202	158	-44	86
10	57N	44E	217	242	198	-44	58
11	54N	46E	154	188	120	-68	80

The similarity of shape to the average of the observed trends is demonstrated in Figure 8. E1 in particular shows very good agreement, with the mean excesses above the summer minimum for the months with all four stations available, with no differences exceeding 13mm. The lack of a peak associated with spring snowmelt in the observed is not a general feature of the V&Y data, several stations having a mean end of month value more than 20mm above the adjacent months, in two cases (near 54N, 70-90E) about 40mm above. Such a peak is also noted by Robock et al. (1995), in their analysis of Russian soil moisture data. Possibly it is missing because it does not occur near the monthly observation date (28th) at the two stations with full monthly data.

The similarity between modelled and observed is perhaps surprising in view of the large differences in precipitation and evaporation; evidently these roughly cancel through the summer as runoff from the top metre is relatively small at this time compared with the other terms. In the experiments, this cancellation follows from the low rainfall and exhaustion of the soil water store - in such arid climates, summer evaporation is obviously constrained by current rainfall. The similarity of precipitation and evaporation for the real data suggests that the opposite control may be operating - restriction of rainfall by evaporation. Such a constraint has been

found to be important in a number of model experiments (e.g. Rowntree and Bolton, 1983; Mintz, 1984). Recently, Milly and Dunne (1994) showed the importance, because of this mechanism, of correct specification of the water-holding capacity of the plant-root zone. In their experiments, part of the increased evaporation resulting from the greater water store was used to increase the precipitation, part to reduce the convergence of water from the ocean, the relative increase in each varying geographically.

Causes of limited water capacity

An obvious implication of the comparison between the modelled and observed soil moistures is that the HC model has too small water-holding capacity in the region we are considering. This version of the model has a single layer, depth equal to the estimated root depth D for the vegetation type specified for that point. The drainage out of the layer is

$$K = K_S \, s^c \qquad (3)$$

where $s = (\theta - \theta_w)/(\theta_S - \theta_w)$; here θ is the available soil moisture and θs and θ_w the saturation and wilting values of θ, K_S is the saturation value of the hydraulic conductivity and c is an exponent. As discussed by Warrilow et al. (1986) who based this fit on a review of the literature, there is a wide range of values for these soil characteristics reported in the literature, this large range no doubt being occasioned partly by the difficulty of fitting profiles such as (3) to the observed data.

Likely reasons for the underestimation of water-holding capacity are:

- too shallow a root zone;
- neglect of the transfer of water from below the root zone during summer;
- incorrect specification of the soil hydrological parameters - specifically, too low a value of $(\theta - \theta_w)$ at the point where drainage effectively ceases.

These are considered in turn below.

Root depth: The root depth in the model for the 55°N point is 0.67m, the specified value in the model for the arable cropland and pasture vegetation prescribed for this point in the model. For the natural forest vegetation of the zone in which V&Y state that these stations lie, a deeper root depth of about 0.9m would be prescribed; the vegetation of the actual plots where the observations were made is grass (Vinnikov, personal communication), as for the plots considered by Robock et al. (1995). Robock et al. comment that preliminary results for one site suggest little difference in soil moisture between forest- and grass-covered plots. For the stations they analysed, both the soil moisture and

its seasonal variation in the 50-100cm layer were generally less than that in the 0-50cm layer, suggesting that the roots are mainly drawing water from the upper 50cm.

Water transfer from below the root zone: The lack of a gradient to transfer water upwards noted above for the observations suggests that the neglect of water below the root zone cannot be the basic cause of the model's aridity; however, as the model's root zone does dry out, it could still contribute to it. However, an experiment with a new multi-layer version of the model in which such transfer could be represented gave only minor improvements in the simulation; upward transfers into the root zone mostly reached only about 0.1 mm/day even in the driest period.

Soil hydrological properties -
(i) specification of soil types: The third possibility to consider is that of the soil hydrological properties. There are a number of aspects to this problem. The first is the specification of the soil types in the model. Wilson and Henderson-Sellers (1985) define the soils for most of this region as of "medium" texture. Their soil texture triangle shows that the effective criteria for this are a clay content of less than about 35%, and a sand content of less than about 65%. Thus there is a large range of sand content. Cosby et al. (1984)'s statistical analysis of 11 of the 12 soils in the USDA soils textural triangle, showed that the saturated hydraulic conductivity displayed greatest sensitivity to sand content, with a smaller but still substantial dependence on clay content, such that K_S increased with increasing sand and decreasing clay content. To assess the importance of the consequent uncertainty, the "field capacities", defined by a drainage rate of 0.01 mm/hour or about 7.2 mm/month, have been analysed for two soils at opposite extremes of this "medium" range. The names and definitions of these (from Cosby et al. (1984)) are given in Table 4.

Table 4: Soil types used for analysis of soil hydrological characteristics, with contents of silt, sand and clay from Cosby et al. (1984)

Class	% silt	% sand	% clay
(1) Silty clay loam	56	10	34
(2) Sandy loam	32	58	10

(ii) Definition of soil characteristics - Clapp-Hornberger: The second aspect of the problem is the different characteristics deduced for the soils in different analyses. These analyses have typically been in terms of the Clapp-Hornberger equations.

Hydraulic conductivity K, to which, in order to limit the complexity of the problem, this analysis is restricted, is then derived using the simple exponential relation

$$K = K_S \, (\theta/\theta_S)^c \qquad (4)$$

where θ is the total soil moisture content, and θ_S is its saturation value; c is an exponent, which is generally assumed to be related to the slope b of the retention curve by a relation of form

$$c = 2b + 2 + L \qquad (5)$$

where L is here assumed to be 0.5 (L=1 is used by Clapp and Hornberger (1976), but the lower value is used here for consistency with the Mualem - Van Genuchten relation discussed below - the effect of this on the present analysis is small). Table 5 shows the values of K_S and b recommended by Cosby et al. (1984), together with values for an "available field capacity" for the two soil types. I define this as $(\theta_f - \theta_w)$; here θ_w is the wilting value of θ, the value at which evapotranspiration ceases. Although the retained moisture θ_r, the amount not available for drainage, can be less than θ_w, the rate of drainage at θ_w is typically so small using Equations (4) and (6) that the effective limit to drying out is that set by the wilting point. However, it should be noted that considerable uncertainty is also associated with the value of θ_w: Warrilow et al. (1986) noted a range for loam soil of from 0.05 (Marshall and Holmes, 1979) to 0.135 (Rutter, 1975). Even without taking this into account, Table 5 shows there are large uncertainties in "available field capacity" associated with the broad range of soils covered by the medium classification.

(iii) Alternative definitions of soil characteristics: A further uncertainty in "available field capacity" as we have defined it stems from the use of different relations to derive K. Van Genuchten et al. (1991) have studied the application of the Mualem - Van Genuchten relation (Van Genuchten, 1980) to the same soils discussed above. According to this relation,

$$K = K_S \, S^L \, [1 - \{1 - S^{1/m}\}^m]^2 \qquad (6)$$

where $S = (\theta - \theta_r)/(\theta_S - \theta_r)$. In terms of the parameters used in Equ. (4, 5),

$$m = 1/(1+b) \qquad (7)$$

is assumed in calculations for Table 5. Van Genuchten et al. (1991) considered two analyses of the soil characteristics, those due to Carsel and Parrish (1988) and Rawls et al. (1982). Data for each of these are included, with b specified according to (7).

Table 5: Soil characteristics for soils (1) and (2) as defined in Table 4 according to different analyses (Cosby et al. (1984); Carsel and Parrish (1988); Rawls et al. (1982)) (see text for details) - also for medium soils as used in the HC model.

Analysis	Soil	K_s m/day	b	θ_r	θ_s	"S_f"	θ_f	$\theta_f-\theta_w$
Cosby	(1)	.176	8.72	-	.464	.718	.333	.228
	(2)	.452	4.74	-	.434	.532	.231	.126
Carsel	(1)	.0168	4.35	.089	.43	.88	.389	.284
	(2)	1.061	1.123	.065	.41	.234	.146	.041
Rawls	(1)	.036	6.62	.040	.432	.91	.317	.212
	(2)	.622	3.11	.041	.412	.561	.249	.144
Warrilow	Med.	.156	2.75	.105	.46	.445	.263	.158

Notes: S_f is S as defined in text for $\theta = \theta_f$ for the Carsel and Parrish (1988) and Rawls et al. (1982) analyses and for the Warrilow scheme; for Cosby et al. (1984), it is θ_f/θ_s.

The values in the last columns under $\theta_f-\theta_w$ are a measure of the amount of water available for evaporation at the end of the wet season, when evaporation becomes the dominant term in the soil water balance. There is clearly a large uncertainty depending not only on where in the WHS "medium" range the soil lies, but also on which particular formulation is used to relate the soil hydraulic conductivity to the soil moisture and also on the value assumed for θ_w or θ_r. Note that in the above we have simplified the problem by considering a single-layer soil, so allowing omission of the soil suction or pressure head from consideration. The values used by V&Y for the region we are considering are well within the range for "medium" soils. The Warrilow model values are near the lower end of the range, especially after allowing for θ_w.

Soil water balance

Observed using runoff data: To expand this assessment of how the observed data may be used to validate the model's simulation, we need to study the soil water balance. We consider first the real case. Figure 9 summarizes the water balance data from Korzun et al. (1978) for Moscow for comparison with the soil moisture using Eq. (2).

The residuals (P-E-Y) should not be interpreted as changes in soil moisture alone but rather as changes in the sum of soil moisture, ground water and snow water content. From October to March they are likely to be divided between moistening of the soil and accumulation

of snow mass. Snow in the Moscow area typically melts around the start of April and covers the ground during late November as shown by ground-based data (Arctic Construction and Frost Effects Laboratory, 1954; Lydolph, 1977) and satellite-based data (Matson et al., 1986), so snow should not affect the data from May to October. The total drying due to excess of evaporation and runoff over precipitation accumulated for May to July inclusive is about 170mm. We may compare this with the change (Table 3) in soil moisture from late April to late July in the top metre at soil moisture stations near Moscow. The observed drying falls short by from 80 to 130mm.

This may be because the water running off in the rivers is coming not from the top metre of soil but from ground water. This would be consistent with the observations that, at stations 10 and 11 by 28 May and at station 8 soon after, soil moisture is below "field capacity" - which V&Y define as "the maximum water content which can be kept in the soil by the forces counteracting gravitation when there is no contact with groundwater". Runoff in May-July accounts for nearly 60mm, so that a further contribution of 20-70mm (depending on which station in Table 3 one uses) is required. The evaporation is clearly uncertain and with a total for the three months of 317mm, a 10% overestimate could be sufficient to explain the discrepancy and make (P-E) consistent with the soil moisture decreases.

Observed using soil moisture: Another approach to this problem is to use the soil moisture data for 1972-85 together with the Moscow precipitation data for the same period; we continue to use the evaporation data from Korzun et al. (1978) in the absence of an alternative. Equ. (1) shows that this allows us to deduce the sum of the runoff (Y) and the change in snowmass (d(Sn)/dt) as:

$$P - E - d(SM)/dt = Y + d(Sn)/dt$$

We must remember that Y is the loss of water to groundwater as well as the loss direct to rivers, so that even in summer the result is not expected to be equal to the river runoff. Figure 10 shows the components of the budget. The soil moisture data are a simple mean of the data for the four stations in Table 3 where available (see Fig. 7), and a best estimate based on the available data otherwise.

The residual can be interpreted as dominated by accumulation and loss of snowmass from November to April inclusive and as loss of water below 1 metre depth to rivers from May to July. Essentially, the winter precipitation appears aas river runoff in spring and early summer. The residual is small from August to

October.

Comparison with model: Finally, Figure 11 compares the above estimates of (P-E-d(SM)) for reality with similar data from the model experiments. The most important differences occur in spring; for E1, this may just be due to timing of snowmelt - the sum for March and April is the same as for the observed data; for E2, the greater precipitation and lesser evaporation appear responsible. The discrepancy in November appears to be due to the model's soil absorbing more of the precipitation than is observed and so reaching its winter plateau sooner.

Discussion and conclusions

The observed data are inadequate in several respects. The annual precipitation, evaporation and runoff do not balance in available analyses; the balance could apparently be improved by reducing the corrections for exposure applied to the precipiation data but, as the evaporation data are particularly uncertain, this is not necessarily an appropriate solution. Studies of the annual water balance should be made in catchments with large winter snowfall during field experiments in which evaporation is measured (e.g. GCIP, BOREAS and Mackenzie basin experiments)

Despite these shortcomings, the observations, particularly those of soil moisture, indicate clear problems with the model simulations. Soil water-holding capacity appears insufficient in the HC model. Use of soil texture categories which encompass too wide a range of soils may contribute to this. Even for a given soil, a wide range of available water-holding capacities may be calculated depending on the choices made from the literature for the soil characteristics including the equations used to relate them to soil moisture and for the moisture remaining at wilting point. This wide range needs to be reduced by further analysis of soil data designed to give a realistic interpretation of the soil observations for the soil moisture values of most importance in determining the effective water-holding capacities of the soil layers in those layers from which water is drawn for evaporation.

These problems are not easily solved. Soil characteristics are not directly observable from space, so that construction of global data sets must rely either on collation and analysis of existing ground-based data together with further extensive observations in regions with insufficient data, or on the use of indirect methods. The latter may attempt to choose those soil and surface characteristics which provide the best fit to relatively well-observed data such as river runoff and/or remotely-sensed surface temperatures (Mahfouf, 1991).

Figure captions

Figure 1: Precipitation for June and August as estimated from observations (a, d) by Legates and Willmott (1990) and simulated by the HC model in experiment E1 (b, e) and E2 (c, f). Contours at 0.1, 0.5, 1, 2, 5 mm/day; stippled above 2 mm/day.

Figure 2: Soil moisture (mm) near end of month (1 = January, 2 = February, etc.) as simulated at 57.5° 37.5°E for Experiments E1 and E2, and as observed (average of stations 5 and 10 from V&Y - about 58°N 28°E, 57°N 44°E).

Figure 3: As Fig. 2 for model at 50°N 48.75°E and observed (average of points 33 and 35 from V&Y - about 52°N 51°E and 49°N 46°E).

Figure 4: Precipitation (mm/day) as simulated at 55°N 41.25°E for Experiments E1 and E2, and as observed for near Moscow (55.7°N 37.6°E) as gauged (1921-1990) and as estimated from Korzun et al. (1978).

Figure 5: Downward solar radiation (W/m^2) for model at 55°N 41.25°E and for observed near Moscow from Budyko (1963).

Figure 6: Surface evaporation (mm/day) for model and as observed for near Moscow from Korzun et al. (1978) (WWB) and Budyko (1963).

Figure 7: Average soil moistures (mm) near the end of each month for the model and for the 4 stations from V&Y listed in Table 3.

Figure 8: Average soil moistures (mm) near the end of each month relative to the summer minimum for the experiments and for the mean of the 4 stations from V&Y listed in Table 3.

Figure 9: Estimates of water balance components (mm/day) for Moscow from Korzun et al. (1978). (P-E-Y) is the sum of the curves for Precipitation (P), -Evaporation (-E) and -Runoff (-Y).

Figure 10: Estimates of water balance components (mm/day) using data for 1972-1985 on soil moisture as in Figure 8 and on precipitation for Moscow, and estimates of longterm mean from Korzun et al. (1978) for evaporation and runoff.

Figure 11: Estimates of loss of water from the top soil layer (P-E-d(SM)) (mm/day) for Moscow and for the model experiments E1 and E2 at 55°N 41.25°E.

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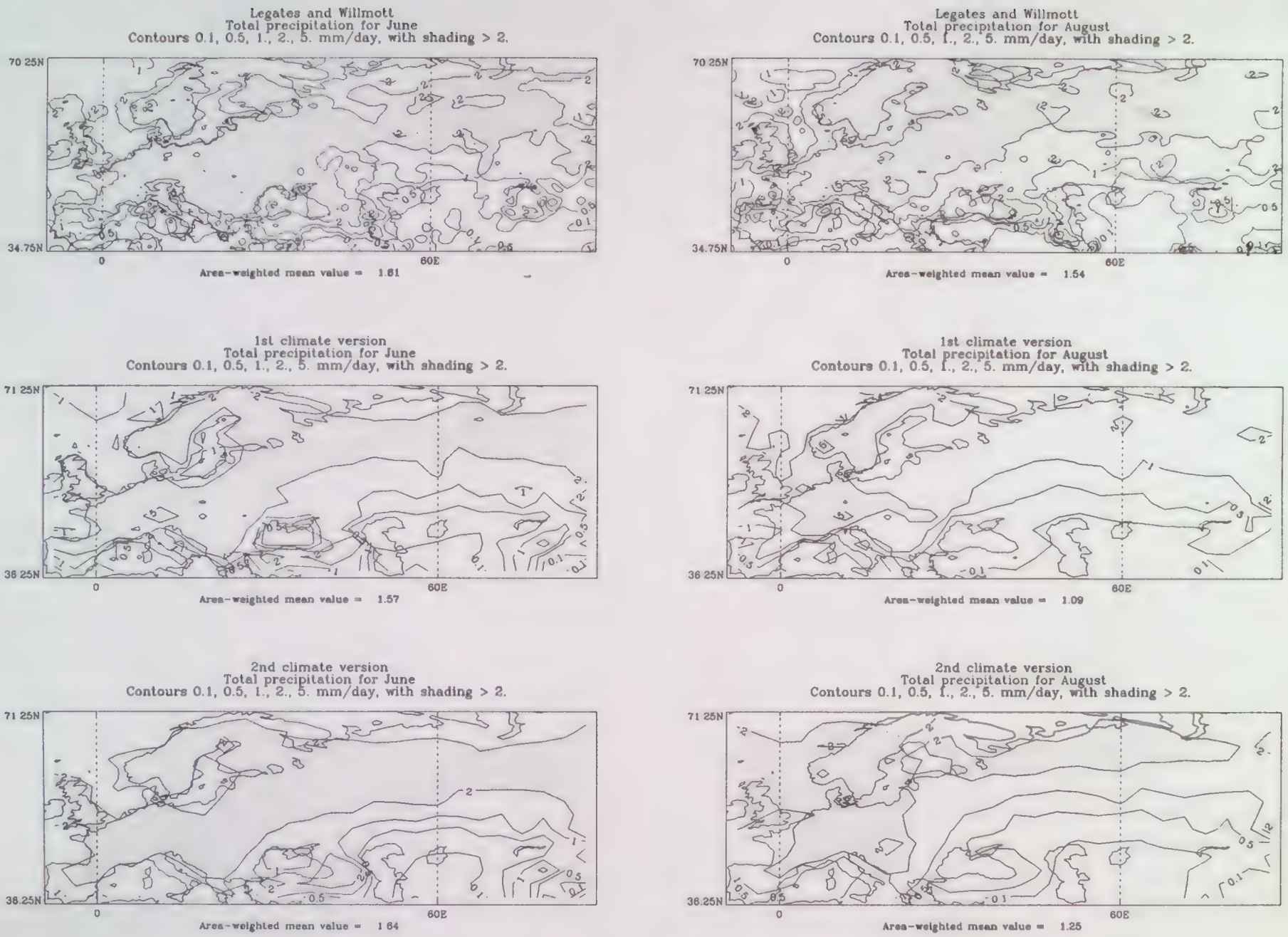
Legates and Willmott
Total precipitation for June
Contours 0.1, 0.5, 1., 2., 5. mm/day, with shading > 2.
70 25N
34.75N
0
60E
Area-weighted mean value = 1.61
Legates and Willmott
Total precipitation for August
Contours 0.1, 0.5, 1., 2., 5. mm/day, with shading > 2.
70 25N
34.75N
0
60E
Area-weighted mean value = 1.54
1st climate version
Total precipitation for June
Contours 0.1, 0.5, 1., 2., 5. mm/day, with shading > 2.
71 25N
36 25N
0
60E
Area-weighted mean value = 1.57
1st climate version
Total precipitation for August
Contours 0.1, 0.5, 1., 2., 5. mm/day, with shading > 2.
71 25N
36 25N
0
60E
Area-weighted mean value = 1.09
2nd climate version
Total precipitation for June
Contours 0.1, 0.5, 1., 2., 5. mm/day, with shading > 2.
71 25N
36.25N
0
60E
Area-weighted mean value = 1 64
2nd climate version
Total precipitation for August
Contours 0.1, 0.5, 1., 2., 5. mm/day, with shading > 2.
71 25N
36.25N
0
60E
Area-weighted mean value = 1.25

Figure 1

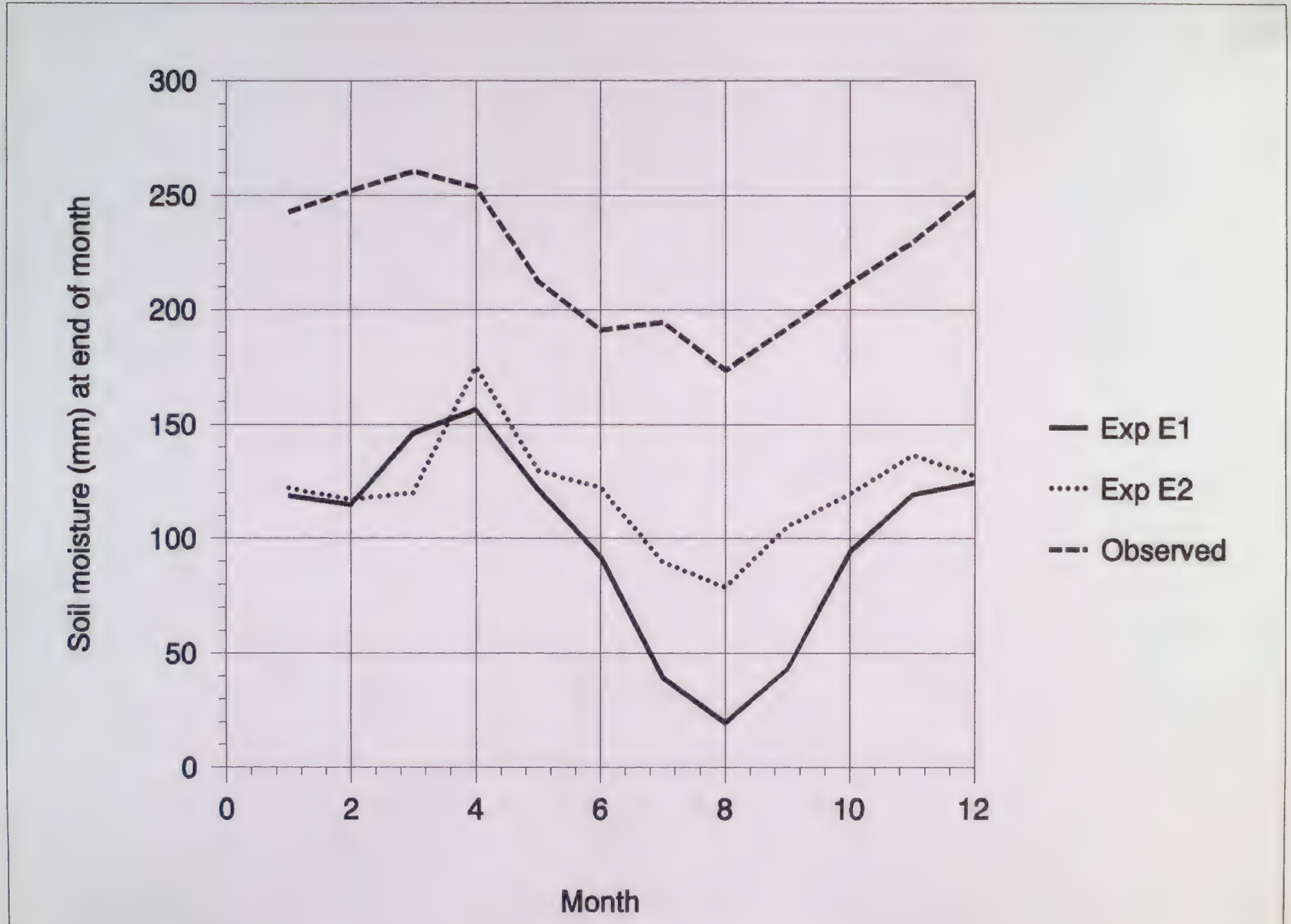
Soil moisture (mm) at end of month
300
250
200
150
100
50
0
0
2
4
6
8
10
12
Month
Exp E1
Exp E2
Observed

Figure 2

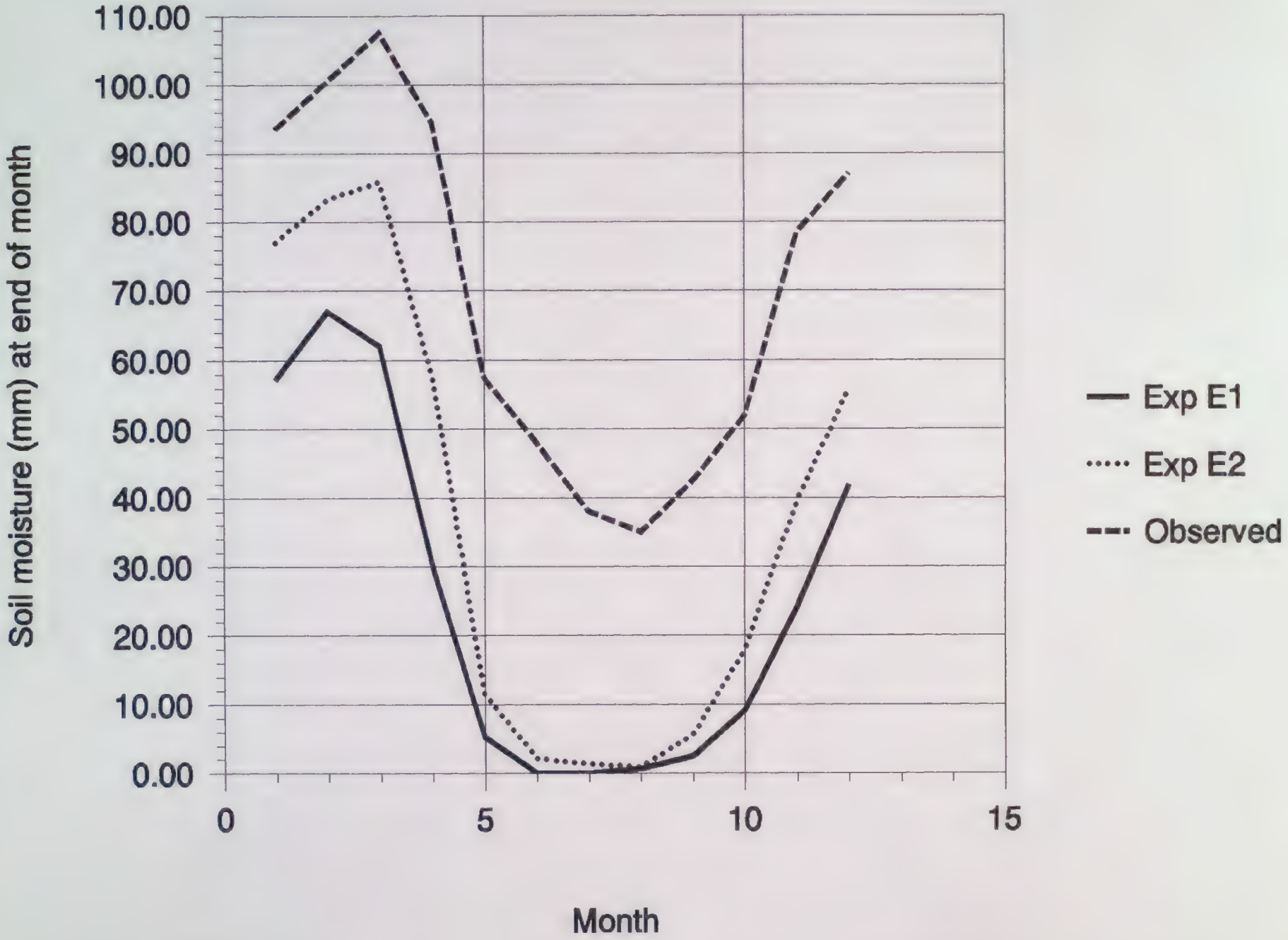
110.00
100.00
90.00
80.00
70.00
60.00
50.00
40.00
30.00
20.00
10.00
0.00
Soil moisture (mm) at end of month
0
5
10
15
Month
Exp E1
Exp E2
Observed

Figure 3

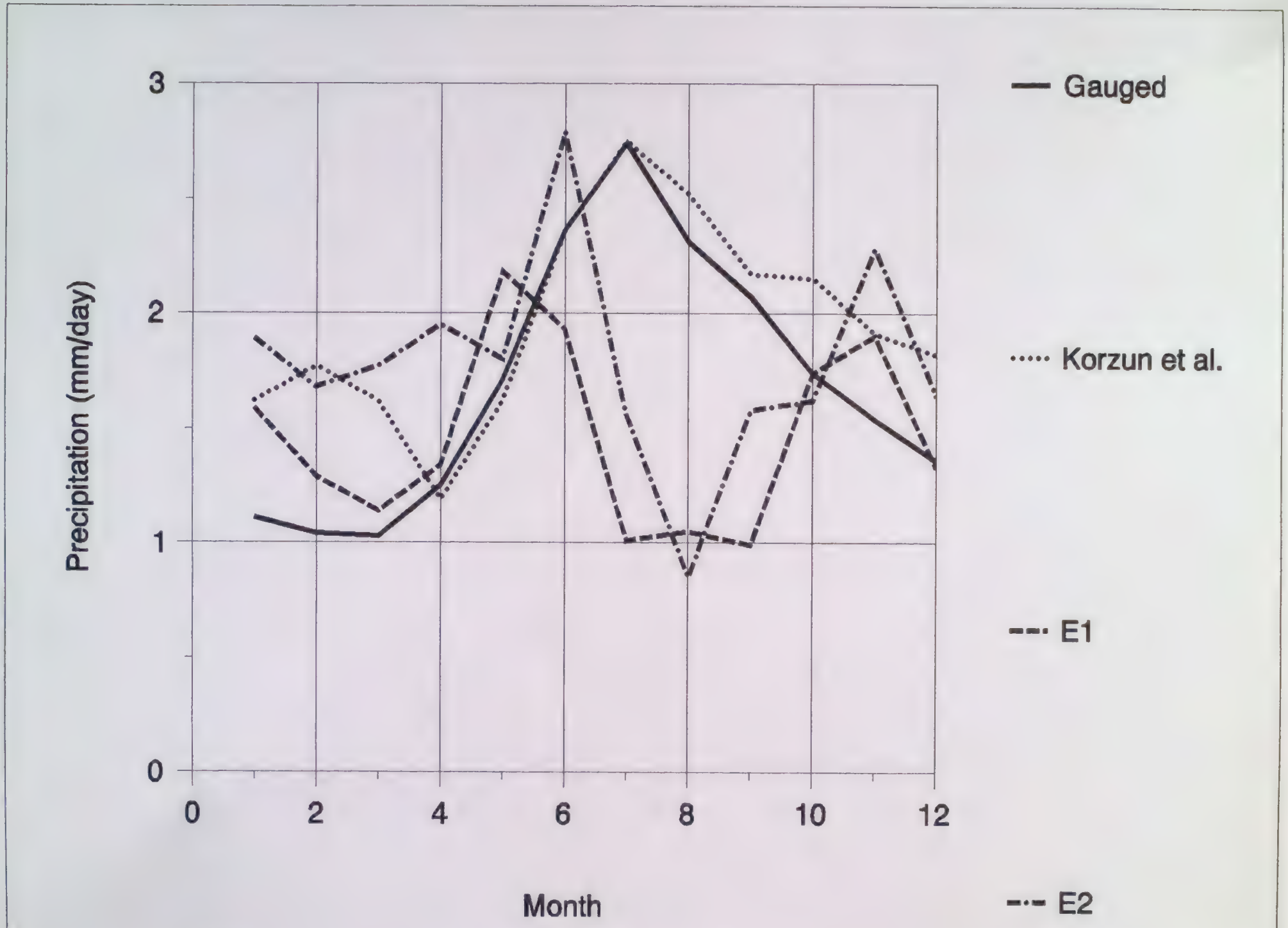
Precipitation (mm/day)
3
2
1
0
0
2
4
6
8
10
12
Month
Gauged
Korzun et al.
E1
E2

Figure 4

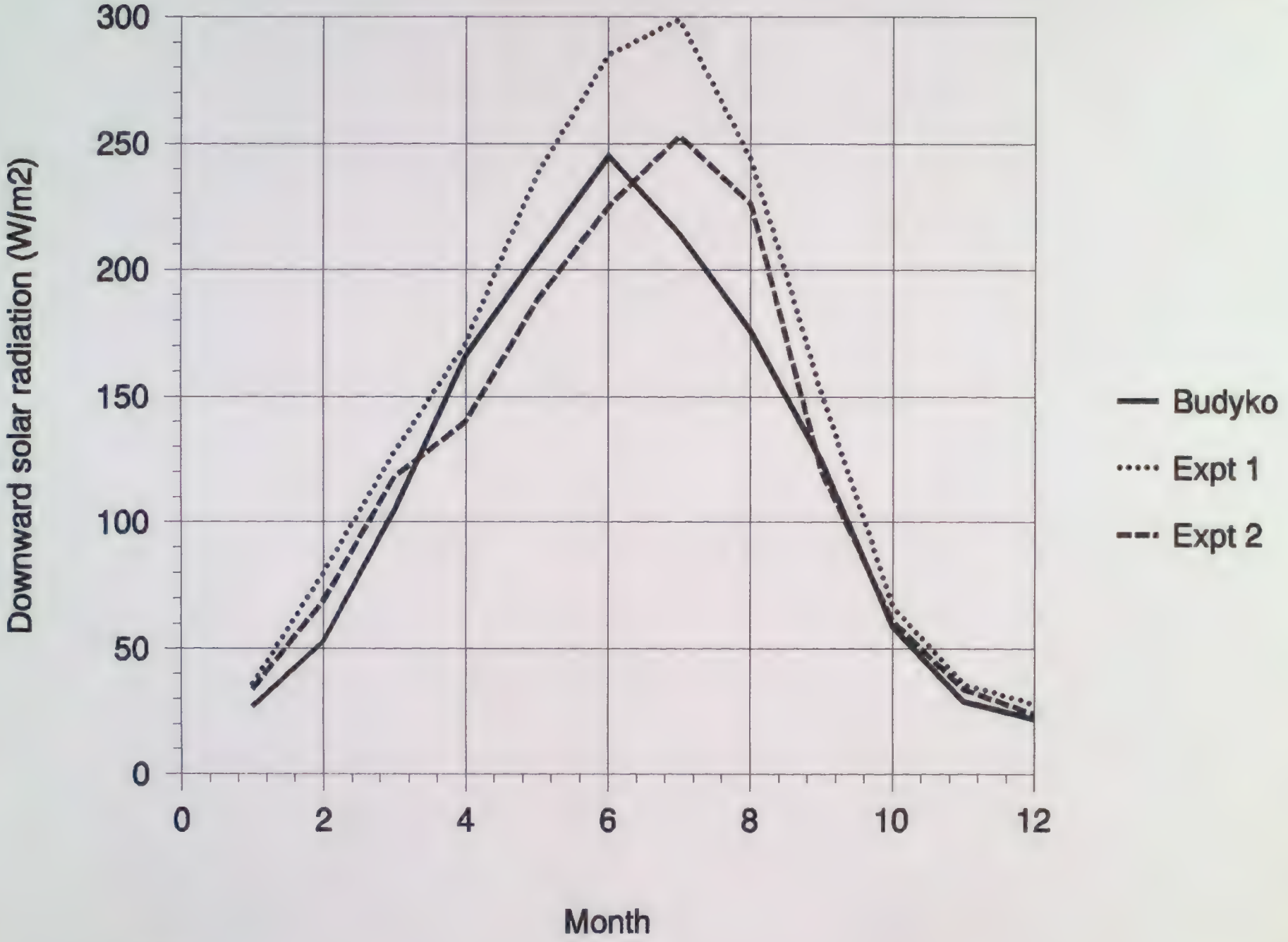
Downward solar radiation (W/m2)
300
250
200
150
100
50
0
0
2
4
6
8
10
12
Month
Budyko
Expt 1
Expt 2

Figure 5

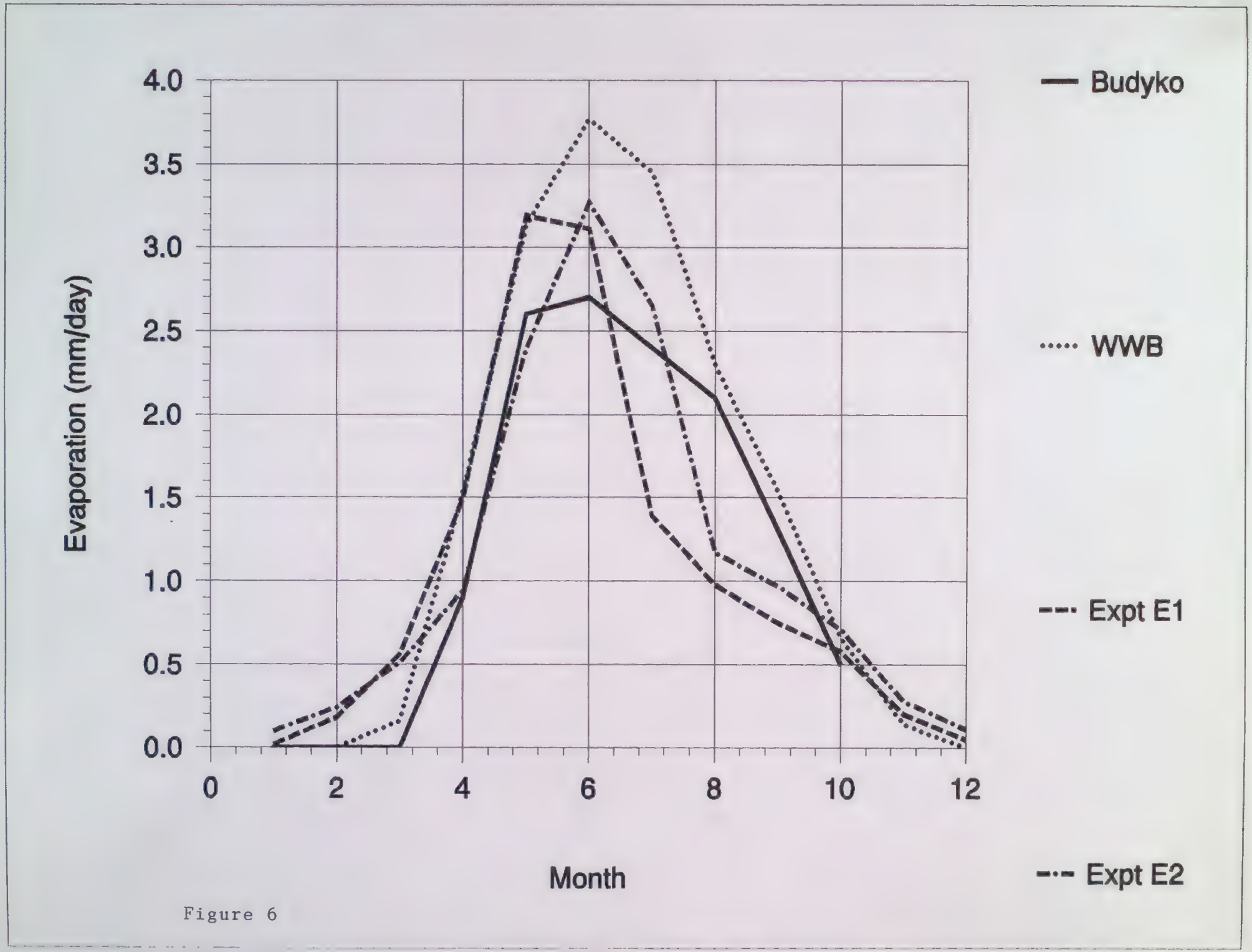
Evaporation (mm/day)
4.0
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0.0
0
2
4
6
8
10
12
Month
Budyko
WWB
Expt E1
Expt E2

Figure 6

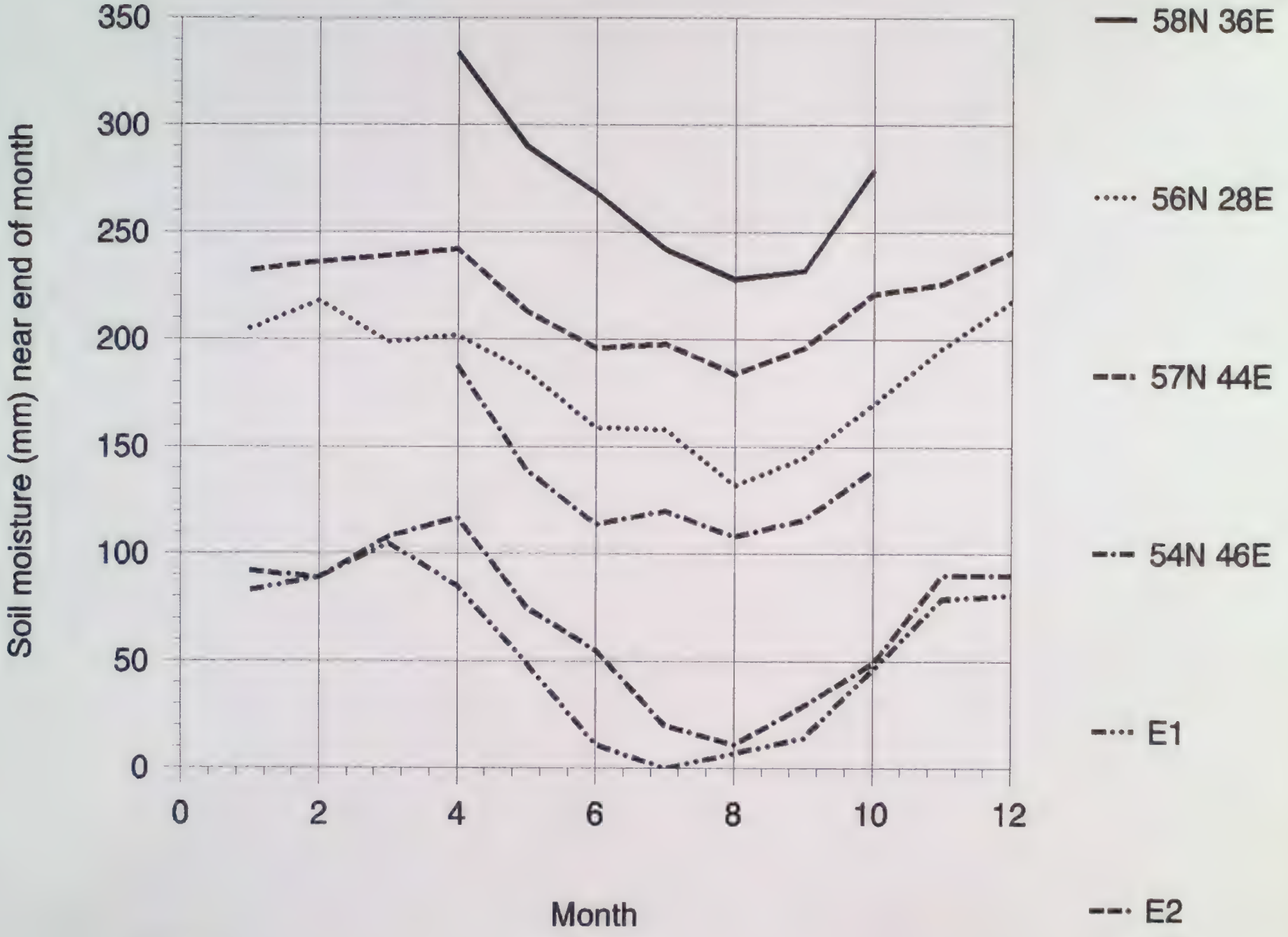
Soil moisture (mm) near end of month
350
300
250
200
150
100
50
0
0
2
4
6
8
10
12
Month
58N 36E
56N 28E
57N 44E
54N 46E
E1
E2

Figure 7

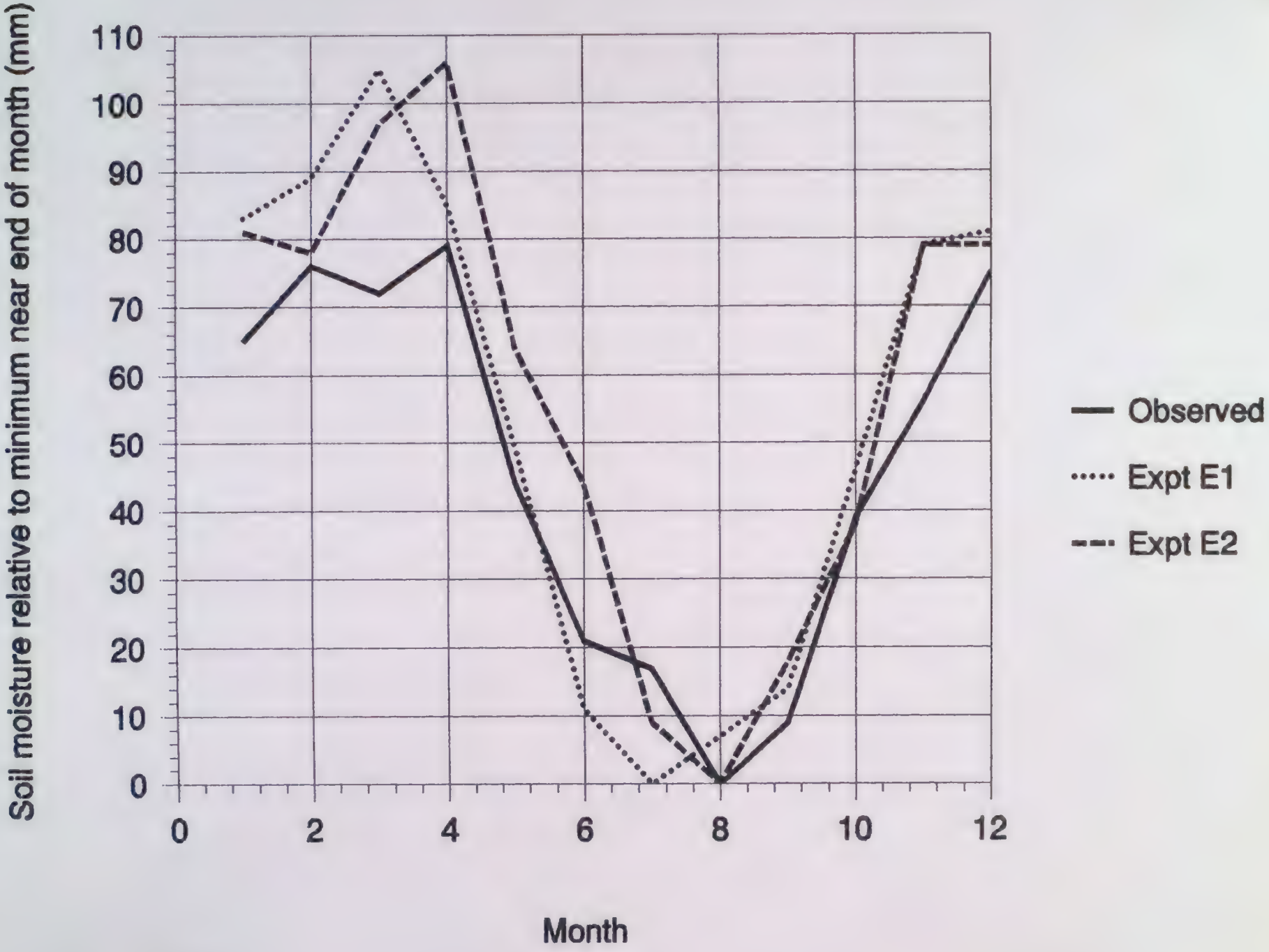
Soil moisture relative to minimum near end of month (mm)
110
100
90
80
70
60
50
40
30
20
10
0
0
2
4
6
8
10
12
Month
Observed
Expt E1
Expt E2

Figure 8

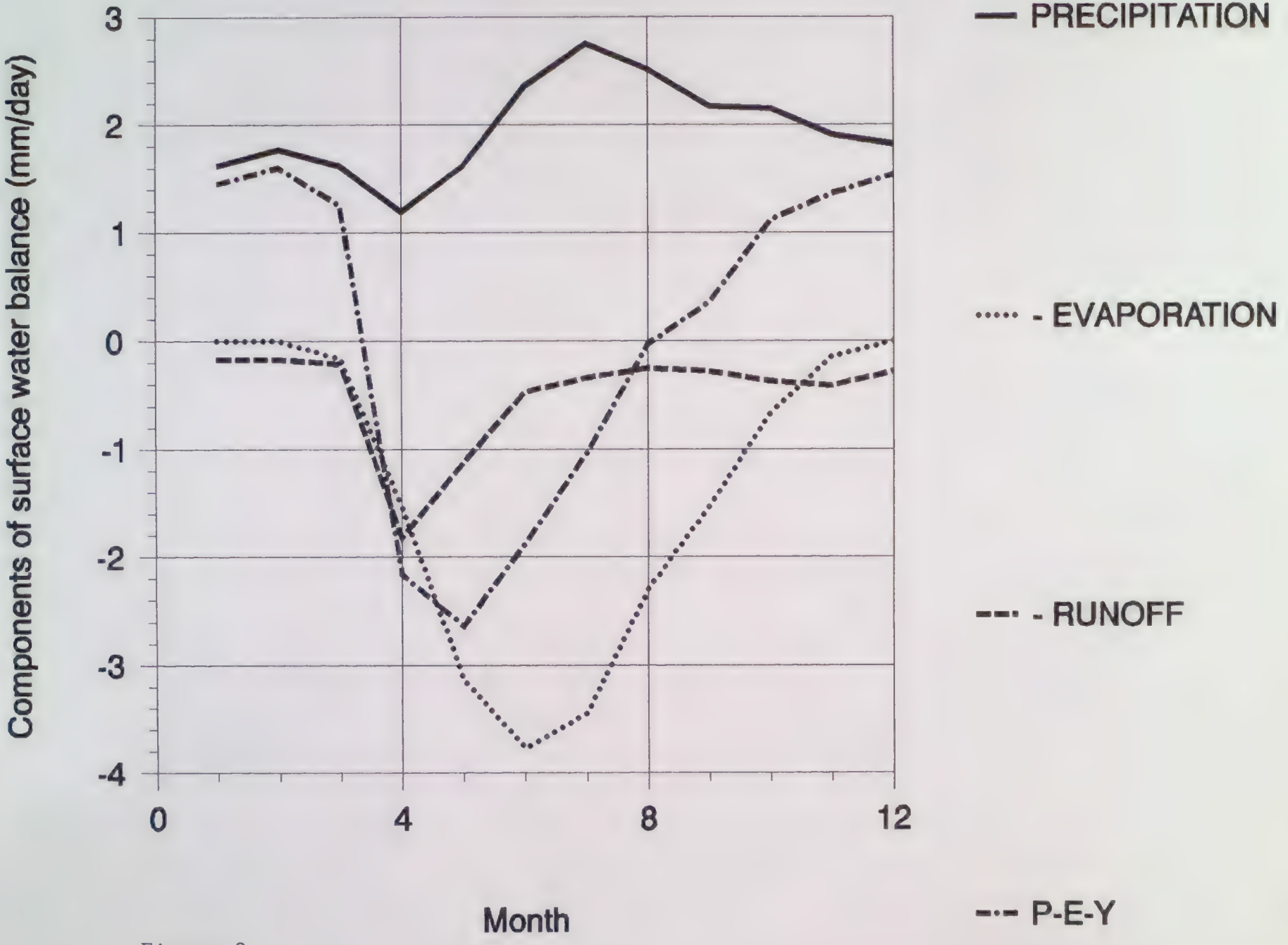
Components of surface water balance (mm/day)
3
2
1
0
-1
-2
-3
-4
0
4
8
12
Month
PRECIPITATION
- EVAPORATION
- RUNOFF
P-E-Y

Figure 9

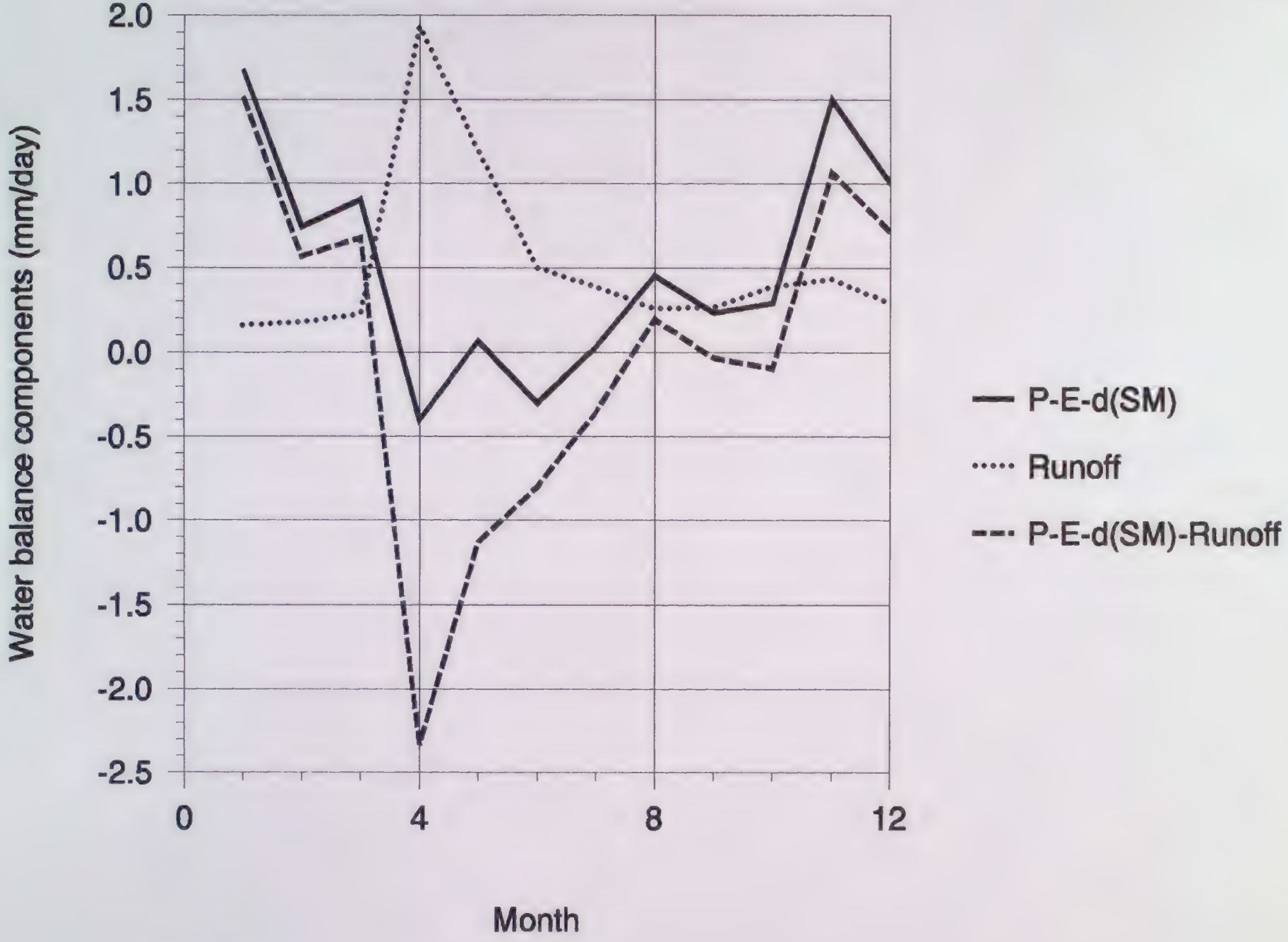
Water balance components (mm/day)
2.0
1.5
1.0
0.5
0.0
-0.5
-1.0
-1.5
-2.0
-2.5
0
4
8
12
Month
P-E-d(SM)
Runoff
P-E-d(SM)-Runoff

Figure 10

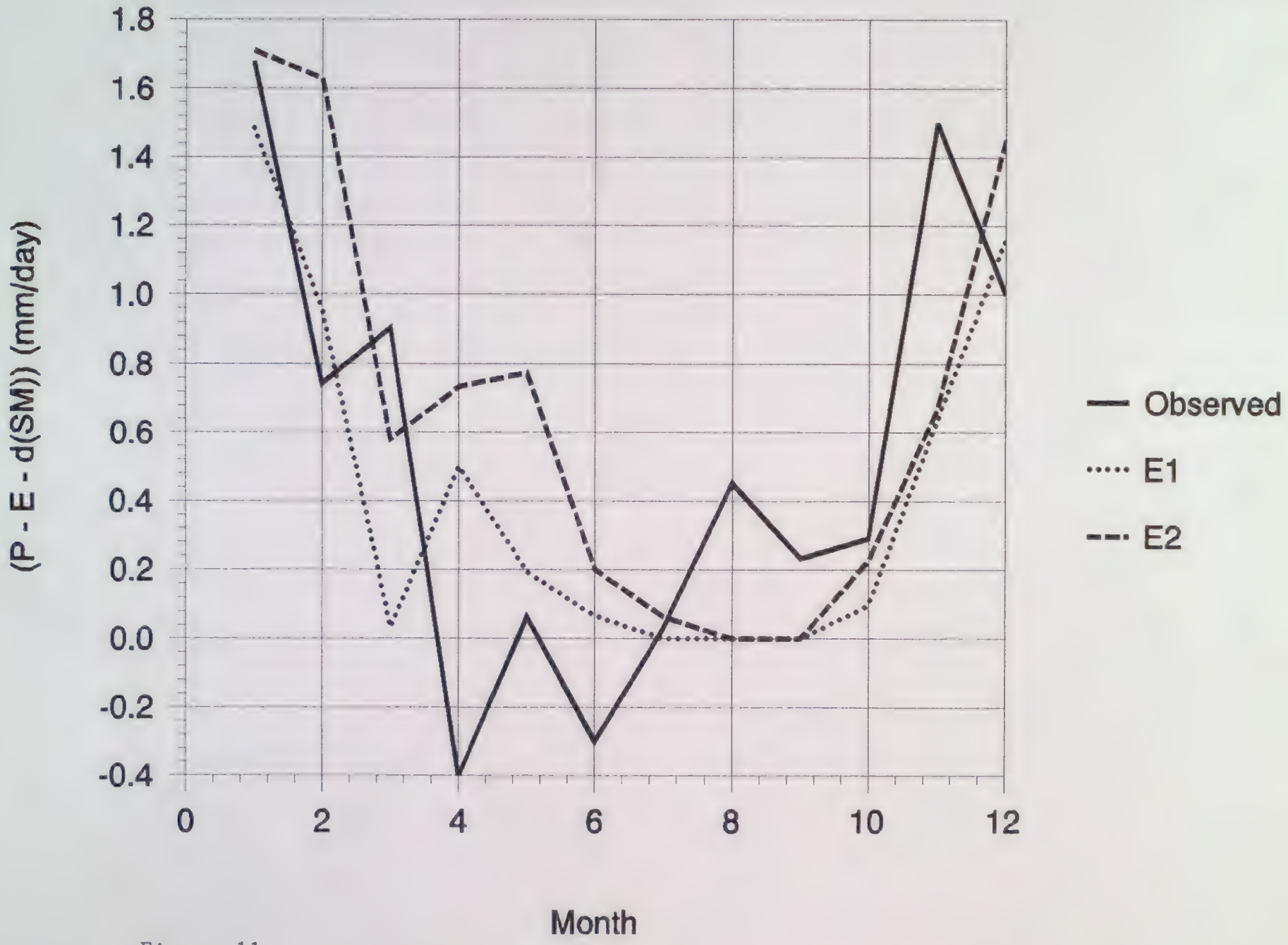
1.8
1.6
1.4
1.2
1.0
0.8
0.6
0.4
0.2
0.0
-0.2
-0.4
0
2
4
6
8
10
12
(P - E - d(SM)) (mm/day)
Month
Observed
E1
E2

Figure 11

CLIMATE RESEARCH TECHNICAL NOTES

CRTN 1	Oct 1990	Estimates of the sensitivity of climate to vegetation changes using the Penman-Monteith equation. P R Rowntree
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CRTN 4	Oct 1990	Low frequency variability of the oceans. C K Folland, A Colman, D E Parker and A Bevan
CRTN 5	Dec 1990	A comparison of 11-level General Circulation Model Simulations with observations in the East Sahel. K Maskell
CRTN 6	Dec 1990	Climate Change Prediction. J F B Mitchell and Qing-cun Zeng
CRTN 7	Jan 1991	Deforestation of Amazonia - modelling the effects of albedo change. M F Mylne and P R Rowntree
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CLIMATE RESEARCH TECHNICAL NOTES

CRTN 13	Jul 1991	Modelling changes in climate due to enhanced CO2, the role of atmospheric dynamics, cloud and moisture. C A Senior, J F B Mitchell, H Le Treut and Z-X Li
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CRTN 19	Sep 1991	The Oceanic Carbon Cycle. N K Taylor
CRTN 20	Nov 1991	Worldwide ocean-atmosphere surface fields in Sahel wet and dry years using provisionally corrected surface wind data. M N Ward
CRTN 21	Dec 1991	Coupled tropical ocean global atmosphere models at the UKMO. M K Davey, C Gordon, S Ineson and S Lawrence
CRTN 22	Dec 1991	Empirical parameterisation of tropical ocean-atmosphere coupling. M Allen, M K Davey, D L T Anderson and P D Killworth
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CLIMATE RESEARCH TECHNICAL NOTES

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CRTN 28	Jun 1992	Development of worldwide marine data eigenvectors since 1985 A W Colman
CRTN 29	Jun 1992	A tropical ocean model with reduced physics M K Davey
CRTN 30	Jun 1992	International Temperature Workshop, Boulder, Colorado, USA, 16 January 1992 Edited by D E Parker
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CRTN 32	Sep 1992	A prediction of the transient response of climate J M Murphy
CRTN 33	Nov 1992	LEPS scores for assessing climate model simulations and long-range forecasts. C K Folland
CRTN 34	Jan 1993	Stratospheric Data Assimilation System Guide Editer: R Swinbank
CRTN 35	Mar 1993	A Stratosphere-Troposphere Data Assimilation System R Swinbank and A O'Neill
CRTN 36	Mar 1993	Validation of hydrological schemes for climate models against catchment data. P R Rowntree and J Lean

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CRTN 46	Apr 1994	Global data required for monitoring climate change. D E Parker and C K Folland
CRTN 47	May 1994	Seasonal uptake of anthropogenic CO_2 in an ocean general circulation model. N K Taylor

CLIMATE RESEARCH TECHNICAL NOTES

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CRTN 50	Sep 1994	Simulation of global mean temperature using a box-diffusion climate model P R Rowntree
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CRTN 54	Mar 1995	The representation of moist convection in atmospheric models. D Gregory
CRTN 55	Mar 1995	A consistent treatment of the evaporation of rain and snow for use in large-scale models D Gregory
CRTN 56	Apr 1995	Workshop on simulations of the Climate of the Twentieth Century using GISST, 28-30 November 1994, Hadley Centre, Bracknell, UK Edited by C K Folland and D P Rowell
CRTN 57	May 1995	Enhanced shortwave cloud radiative forcing due to anthropogenic aerosols. S E Swartz and A Slingo
CRTN 58	May 1995	The simulation of the tropical oceans in models of different horizontal resolution. C Gordon, D K Wright and C M Roberts
CRTN 59	Sep 1995	The water budget of middle latitude continental regions - a modelling and observational study P R Rowntree